Dynamical Climatology

Climate sensitivity and past climates: evidence from numerical studies

by

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Climate Sensitivity and Past Climates: Evidence from Numerical Studies

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Abstract

The use of three dimensional general circulation models to study past climates is considered. The importance of internal feedbacks (due to changes in water vapour, sea-ice and cloud) is reviewed, and the physical basis of the relevant parametrizations is assessed. It is concluded that at present, models can provide qualitative guidance on large scale features of past climates for periods when the relevant boundary conditions can be specified accurately and are changing sufficiently slowly.

1. Introduction

There is much interest in the possibility that climate can change abruptly (for example Ghazi, 1983). Palaeoclimatic records indicate the transitions between glacial and inter-glacial periods were not gradual, but punctuated with warmer and colder episodes, such as the cooling in the north east Atlantic during the Younger Dryas period at the end of the last ice age. The topic is of more than academic interest, as by the end of the 21st century, atmospheric carbon dioxide concentrations may have doubled above the pre-industrial level, producing a global warming of 1.5 to 4.5°C (for example, US National Acadamy of Sciences, 1983).

The causes of climate change are very much a matter for speculation. There is evidence that variations in the earth's orbit may be responsible for oscillations between glacial and interglacial periods. Major volcanic eruptions such as that of Krakatoa in 1883 deposit large quantities of aerosol in the stratosphere where they may remain in sufficient quantities to reduce the solar radiation reaching the surface for several years. Oscillations may occur due to internal variations generated within the climate system. Anomalies in deep ocean temperature may persist for a century or more; variations in the major ice sheets will occur over much longer timescales.

In numerical studies of climate, it is usual to compare a control simulation with anomaly simulations in which certain boundary conditions (for example continental ice sheets, orbital parameters, sea surface temepratures, surface reflectivities, atmospheric CO_2 concentrations) have been changed. For those not familiar with general circulation models, an account of the construction and use of general circulation models (GCMs) and their use is given by Mitchell (1983a). There have been several 'snap-shot' studies of the last glacial maximum (Manabe and Hahn, 1977; Williams et al, 1974; Gates, 1976; Hansen et al, 1984) in which the effect of prescribing sea surface temperatures and continental ice sheets for 18000 years before present was investigated. Manabe and Broccoli (1985) using a model with a simple representation of the ocean and sea-ice have assessed the impact of changing only the continental ice sheets from their present day configuration to that 18000 years before present. Mitchell (1976, 1977) and more recently Kutzbach and his co-workers (for example,

Kutzbach and Street-Perrott, 1985) have studied the effect of changes in the earth's orbital parameters on climate. This is of particular interest during the Holocene, and especially around 9000 years ago when the seasonal variation of insolation was much more pronounced than at present. Applications of numerical models to the Younger Dryas are discussed by Schneider, Peteet and North (this volume).

One recurring problem in the search for explanations of past climate changes is that the mechanisms proposed do not provide a large enough forcing on their own to explain the observed changes in climate. Thus it seems likely that perturbations may be amplified by the response within the climate system. In this paper, various mechanisms which amplify or damp the response of climate to imposed perturbations are reviewed and assessed, using material from numerical studies of CO_2 and climate. Clearly it is essential to establish that the sensitivity of a model to perturbations is correct if that model is to be used to assess the factors contributing to past climate changes. Recent work has stressed the contribution of feedbacks to globally averaged changes (for example, Dickinson 1985). Here we review internal feedbacks due to changes in water vapour, sea-ice and cloud as deduced from numerical studies of the effect of CO_2, assess their physical significance (that is, are the model feedbacks realistic?) and consider some consequences of the existence of regional variations in climate change.

2. Globally averaged effects

The relative contribution of individual feedback processes has traditionally been analysed using a simple globally averaged energy balance model (for example Hansen et al, 1984; Schlesinger and Mitchell, 1986; Dickinson, 1985). A variable, ΔT, the departure of globally averaged temperature from equilibrium, is given by

$$C \frac{d\Delta T}{dt} + \lambda \Delta T = \Delta Q \qquad (1)$$

where C is the heat capacity of the system, t is time, ΔQ is the imposed external perturbation and λ is the feedback parameter. At equilibrium, $\Delta T = \Delta Q/\lambda$; the smaller λ, the weaker the restoring term on the left hand side of the equation, and the greater the sensitivity of the model (see Appendix). In the absence of internal feedbacks, $\lambda = 3.7$ Wm^{-2} $°C^{-1}$, assuming the unperturbed earth atmosphere system behaves like a black body with an effective emitting temperature of 255 K and has a short wave reflectivity of 30%.

We now consider the principal physical mechanisms known to modify λ, and estimate their magnitude from studies of the effect of increased atmospheric CO_2. Where necessary, we have assumed that doubling atmospheric CO_2 concentrations warms the troposphere/surface system by 4 Wm^{-2} (for example Lal and Ramanathan, 1984), which from the equation (1) gives an equilibrium warming of 1.1 °C in the absence of feedbacks (Table 1).

a. Water Vapour

Increasing temperature by 1K from 288K increases the saturation vapour pressure of water by about 7%. Thus, it is to be expected that the atmospheric water vapour content will increase with temperature. Manabe and Wetherald (1967), on the basis of the observed seasonal variation of atmospheric humidity, concluded that relative humidity, (the percentage of saturation) was approximately constant. This conclusion is supported by numerical studies (as shown in Figure 1, from Mitchell et al, 1985). This being the case, an increase in atmospheric temperature will be accompanied by an enhancement of atmospheric water vapour, increasing the opacity of the atmosphere to long wave radiation and thus enhancing the warming of the atmosphere and surface.

Estimates of the change in λ due to water vapour feedback based on published work are in fair agreement whether from radiative convective models assuming fixed relative humidity ($-1.8\ Wm^{-2}{}^{\circ}C^{-1}$, Manabe and Wetherald 1967; $-1.4\ Wm^{-2}{}^{\circ}C^{-1}$, Hansen et al 1984) or from three dimensional models which generate their own humidity structure ($-1.6\ Wm^{-2}{}^{\circ}C^{-1}$, Mitchell 1983b; $-1.5\ Wm^{-2}{}^{\circ}C^{-1}$, Hansen et al 1984). All the figures quoted are for conditions of average cloudiness and the results from general circulation models take into account the geographical and seasonal variations in atmospheric lapse rate.

b. Sea ice

An increase in atmospheric temperature will tend to reduce highly reflective snow and sea-ice cover, leading to increased absorption of insolation and further warming. Clearly, the strength of this feedback will depend on the areal cover of snow and sea-ice, and hence the temperature of the unperturbed simulation. In general, the colder the control simulation, the greater the sensitivity of the model, as illustrated in Figure 2 (from Spelman and Manabe, 1984). Removal of sea-ice greatly increases the thermal inertia of the surface and reduces the amplitude of the annual cycle - this will enhance a warming in winter and reduce it in summer. Robock (1983), using observational data, estimated the potential effects of changes in snow and ice area, in albedo (due to surface melting) and, in the case of sea-ice, in thermal inertia on climate sensitivity. He deduced a net contribution to λ of $-0.6\ Wm^{-2}\ °C^{-1}$. This is somewhat higher than the -0.3 and $-0.4\ Wm^{-2}\ °C^{-1}$ found by Hansen et al (1984) using two versions of a three dimensional climate model which for present day conditions produced respectively 15% less and 23% more ice cover than observed.

It must be recognised that in all these studies, simple parametrizations of the thermodynamics of sea-ice had been used, and in almost all cases, the advection of sea-ice, and possible changes in ocean salinity and the advection of heat by the ocean have all been neglected. Thus, the above estimates of the strength of sea-ice feedback must be regarded as tentative. Perhaps the most important

points to emerge are that such a feedback will be stronger during glacial rather than interglacial periods, as snow and ice cover are more extensive, and that the associated temperature changes are likely to be greater in winter than in summer.

c. Cloud cover feedback

Clouds reflect solar radiation and absorb and emit infra-red radiation. An increase in cloud cover enhances both the reflection of solar radiation, producing a surface cooling, and the emission of long wave radiation, producing a warming. Current evidence suggests that the former effect is dominant at most latitudes where insolation is significant. The long wave effects may be more important for high, thin cirrus. In addition, an increase in cloud height will also tend to warm the surface, as radiation from the cloud top is emitted to space at a lower temperature.

Simulations with increased CO_2 using three dimensional climate models show a consistent decrease in middle and upper tropospheric cloud at most latitudes, with increases in cloud in the lower stratosphere, particularly in middle and high latitudes (Schlesinger and Mitchell, 1986). Similar changes have been found in an experiment using a version of the Meteorological Office 11-layer model, in which CO_2 amounts were doubled and sea surface temperatures were enhanced by 2°C (Figure 3). The estimated contribution of the cloud changes to λ was about $-0.9\ Wm^{-2}\ °C^{-1}$, which is within the range of values of -1.1 to $-0.8\ Wm^{-2}\ °C^{-1}$ found in the experiments of Hansen et al (1984).

(In the latter experiments the cloud optical thickness was prescribed to decrease with height; since cloud height increased there was an additional positive feedback through the change in optical properties).

It is important that the mechanisms producing the changes in cloudiness in the model simulations are fully understood. The vertical profiles of globally averaged changes in radiative heating, frequency distribution of the level at which moist convection terminates (convective cloud top) and relative humidity from the Meteorological Office experiment are shown in Figure 4. On introducing the globally averaged changes in CO_2, water vapour, temperature and cloud in a one-dimensional version of the radiation scheme used in the experiment, it is found that the large cooling 'spike' near 230 mbs is due mainly to the increase in water vapour, and a similar though slightly less pronounced cooling is found at the same level in an earlier experiment with prescribed cloud. The simulated changes in radiative heating rates make the atmosphere more unstable below 230 mbs, reducing the frequency of detrainment of moist air below that level and increasing it above. Thus relative humidity decreases through most of the troposphere below 230 mbs, and increases above, with associated changes in cloudiness. These changes are consistent with an increase in the height of the tropopause. The observed tropopause is highest in low latitudes and in summer when insolation is strongest: thus there is at least circumstantial observational evidence that the models' response to tropospheric warming is physically correct.

The details of the above changes are obviously sensitive to the parametrization of radiation, particularly with respect to water vapour, and the vertical transport of water. The vertical resolution of most models is poor near the tropopause (generally 3 layers above 250 mb and 4 in the case of the Meteorological Office model) and the vertical extent of cloud is usually limited; for example, in the version of the Meteorological Office model used here it is confined below 200 mb. The parametrization of cloud has to date been based on relative humidity, although attempts have been made to devise a parameterization of cloud liquid water (Charlock and Ramanathan, 1985). This approach has the advantage that cloud optical properties can be related in a consistent manner to the liquid water content (see for example, Stephens, 1978), but as yet, such liquid water content parametrizations are fairly arbitrary.

d. Cloud Optical Properties

The results considered up to this point have assumed that cloud optical properties are invariant with climate. A general increase in atmospheric specific humidity might be expected to lead to an increase in cloud liquid water content and hence the reflectivity of cloud. Stephens (1978) has derived simple parametrizations of the relation between cloud optical properties and cloud liquid water content, based on observational data. He found a gradual increase in cloud reflectivity (albedo) with cloud liquid water content, although the long wave emissivity of thin cloud (for example, cirrus) may also

increase significantly. An increase in albedo would cool the surface; an increase in the long wave emissivity of high cloud would have the opposite effect. It is by no means clear how one can relate cloud liquid water content to large scale model variables (for example temperature, humidity etc). Nevertheless, Somerville and Remer (1984) have attempted to relate cloud liquid water content to temperature, and find that the resulting negative feedback is sufficient to counteract all the positive feedbacks in Table 1! In view of the uncertainties in their parametrization, the magnitude of cloud optical feedback is extremely uncertain, and is a priority for future research.

3. Regional Changes

Since good palaeoclimatic data are generally available over only limited regions, information concerning global mean sensitivity is not particularly useful. Here, we consider three factors which must be borne in mind when comparing model and palaeoclimatic data.

i. Scale. Current general circulation models use a horizontal grid of 250 to 800 km. In practice, because of both the numerical smoothing necessary in all models to ensure computational stability, and the year to year variability inherent in non-linear models (and the real world), it is advisable to consider regions rather than individual grid points. Even then, the regional response of models to a given perturbation is likely to vary considerably. For example, one can conclude from numerical studies that increased atmospheric CO_2 may

produce a drying of the mid-latitude continents in summer (Schlesinger and Mitchell, 1986). Not all models produce such a drying, and amongst those that do, the exact latitudes at which the drying occurs varies. Consequently, it would be unwise to compare observational data over a small region with results from a single numerical study.

ii. Regional Anomalies

Feedbacks act locally as well as globally. Thus, in the experiment using the Meteorological Office model described above, changes in soil moisture and cloud reinforce one another as can be seen over the northern mid-latitude continents in summer (Figure 5). The reduction in soil moisture reduces evaporation and hence the moisture necessary to form cloud; the reduction in cloud increases the energy available for evaporation and enhances the surface drying prevalent at that time of year.

iii. Transient Effects

Because of its smaller thermal inertia, the continental surface will respond more quickly than the ocean to changes in external forcing. As an extreme example, the response of a model in which the ocean has an infinite heat capacity to a 10-fold increase in CO_2 (Figure 6) demonstrates that quite large changes in land-sea temperature contrast can occur. (Doubling CO_2 in the same model produced a response 3 to 4 times smaller).

4. Concluding Remarks

General circulation models can be used to assess the influence of specified boundary conditions on equilibrium climate. Thus, they are a powerful tool for studying periods when the relevant boundary conditions are well known and stationary or, at most, changing on a time scale larger than the response time of the model. They may provide both insight concerning the physical mechanisms leading to climate change and guidance on past climate in regions where palaeo-climatic data are poor or sparse. Even so, it must be recognised that there is some uncertainty in the sensitivity of models to a given perturbation (ie doubling atmospheric CO_2 concentrations) so results are qualitative rather than quantitative. The parametrisation of sea-ice and especially cloud must be more physically based if these uncertainties are to be reduced. Given good widespread palaeo-data from a period in which the boundary conditions are well established, one could validate the sensitivity of climate models directly from observational data. However, until there is both a significant improvement in the quality and coverage of past data and a better agreement in the regional changes produced by models in response to the same perturbation, one cannot have much confidence in such an approach.

5. Acknowledgements

I would like to thank C A Wilson and W M Cunnington for their help in the study of changes in cloud using the Meteorological Office 11-layer model, and P R Rowntree for comments on the manuscript.

6. Appendix - Stability, Sensitivity and Feedbacks

In a recent paper, Hansen et al (1984) made an analysis of feedbacks in a three dimensional climate model, based on certain simplifying assumptions and procedures derived from studies of feedbacks. In this note, we derive some of the results used by Hansen et al using elementary calculus, and clarify the underlying assumptions.

Suppose the rate of change with time $\frac{dT}{dt}$ of a variable T in the climate system (for example, temperature) can be expressed in the form

$$\frac{dT}{dt} = F\,(T, \{X_i\}\ , \{Y_j(T)\}\ ) \qquad (1)$$

where F (for example, heat flux per unit heat capacity of the system) can be expressed as a function of T, parameters X_i assumed to be independent of T (for example, the solar constant, atmospheric CO_2 concentration) and parameters Y_j which are known to be dependent on T (for example, planetary albedo, atmospheric emissivity). Equilibrium values of (1) are given by $\frac{dT}{dt} = 0$, so (1) can be expanded (to first order) about an equilibrium value T_e by

$$\frac{dT}{dt} \simeq \lambda (T - T_e) \qquad (2)$$

where $\lambda = -\frac{dF}{dT}\Big|_{T=T_e}$. If $\lambda > 0$ perturbations on T decay exponentially to zero with e-folding time $1/\lambda$, but if $\lambda < 0$, they grow exponentially and the equilibrium is unstable (see for example, Fraedrich 1978, 1979).

From (1)

$$\lambda = -\left(\frac{\partial F}{\partial T} + \sum_j \frac{\partial F}{\partial Y_j}\frac{dY_j}{dT}\right) \qquad (3)$$

assuming that the Y_j are independent.

Consider small changes ΔT_e in equilibrium T_e due to perturbations ΔX_o in X_o. Since $F\ (T_e, \{X_i\}, \{Y_j\ (T_e)\}\ ) = 0$

$$\frac{dF}{dT}\Delta T_e + \frac{\partial F}{\partial X_o}\Delta X_o = 0 \qquad (4)$$

where $\frac{dF}{dT}$, $\frac{\partial F}{\partial X_o}$ are evaluated at $T = T_e$, whence the sensitivity of T_e to changes in X_o is given by

$$\frac{dT_e}{dX_o} = -\left(1/\frac{dF}{dT}\right)\frac{\partial F}{\partial X_o} = \frac{1}{\lambda}\frac{\partial F}{\partial X_o} \qquad (5)$$

which, with (3) may be rearranged to give

$$\frac{dT_e}{dX_o} = \frac{-\partial F/\partial X_o}{\partial F/\partial T}\left\{\frac{1}{1 + \sum_j \frac{\partial F}{\partial Y_j}\frac{dY_j}{dT}/\frac{\partial F}{\partial T}}\right\} \qquad (6)$$

In Hansen et al's paper, T is the effective radiating temperature of the earth atmosphere system, and F is the net heat flux "into the planet", $\partial F/\partial X_o$ is the rate of change in F due to changing an independent parameter (in their example, the solar constant), $\frac{\partial F}{\partial T}$ is the rate of change in net heating of the planet due to changes in long wave cooling to space ($\frac{\partial F}{\partial T} < 0$), and $-\frac{\partial F/\partial X_o}{\partial F/\partial T}$ gives sensitivity of T_e to changes in X_o in the absence of changes (feedbacks) in the $Y_j(T)$. The amplification due to feedbacks is

$$\frac{1}{1 + \sum_j \frac{\partial F}{\partial Y_j}\frac{dY_j}{dT}/\frac{\partial F}{\partial T}} \equiv \frac{1}{1-g} \qquad (7)$$

where $g = \sum_j g_j$, with $g_j \equiv -\left(\frac{\partial F}{\partial Y_j}\frac{dY_j}{dT}\right)/\frac{\partial F}{\partial T}$

is the "gain" defined in Hansen et al. The expression (6) is valid provided F can be linearized about T_e, and the feedbacks implicit in the $\gamma_j(T)$ are independent. Further estimates of λ and feedback factors are given by Watts (1980) and Dickinson (1985).

Table 1. Typical values of internal feedbacks

Feedback	Contribution to feedback factor ($Wm^{-2}°C^{-1}$)	Feedback factor λ (Cumulative, $Wm^{-2}°C^{-1}$)	Change in surface temperature, °C ($\Delta Q = 4\ Wm^{-2}$)
None		3.7	1.1
Water Vapour	-1.5	2.2	1.7
Sea ice/snow	-0.4	1.8	2.2
Cloud cover	-0.9	0.9	4.4

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Figure Captions

1. Globally averaged annual mean changes in specific humidity (solid line, percentage of unperturbed value) and relative humidity (dashed line, percentage of saturation) simulated in a GCM experiment in which CO_2 was doubled and sea surface temperatures were increased by 2°C (see Mitchell et al, 1985). The vertical co-ordinate is σ (= pressure/surface pressure).

2. Relationship between model's sensitivity to a quadrupling of CO_2 and the simulated control temperature (from Spelman and Manabe, 1984).

3. Zonally averaged changes in cloud (June to August) due to doubling atmospheric CO_2 and enhancing sea surface temperatures (see also Figure 4). Contours every 2% of total cover, areas of increase are shaded.

4. Vertical distribution of globally averaged changes due doubling atmospheric CO_2 concentrations and incresing sea surface temperatures by 2°C in a version of the Meteorological Office 11-layer model with interactive cloud. Results are averaged over two summers (June to August) from a seasonal integration. The vertical co-ordinate is σ (= pressure/surface pressure).

   a. Net radiative heating (K/day)

   b. Frequency distribution of level of termination of moist convection (cumulus top) (percentage of total time).

c. Relative humidity (percentage of saturation).

5. Regional changes in summer (June to August) due to doubling atmospheric CO_2 and enhancing sea surface temperatures by 2°C in a version of the Meteorological Office 11-layer model (see also Figure 3).

a. Soil moisture. Contours every 2 cms, regions of decrese are shaded.

b. Total cloud. Contours every 1% of total cover, areas of decrease are shaded.

6. Changes in continental surface temperature due to increasing atmospheric CO_2 ten-fold, holding sea surface temperatures unchanged (June to August). Contours every 2°C, areas of decrease are shaded. Additional contours at -1, 1, 3.

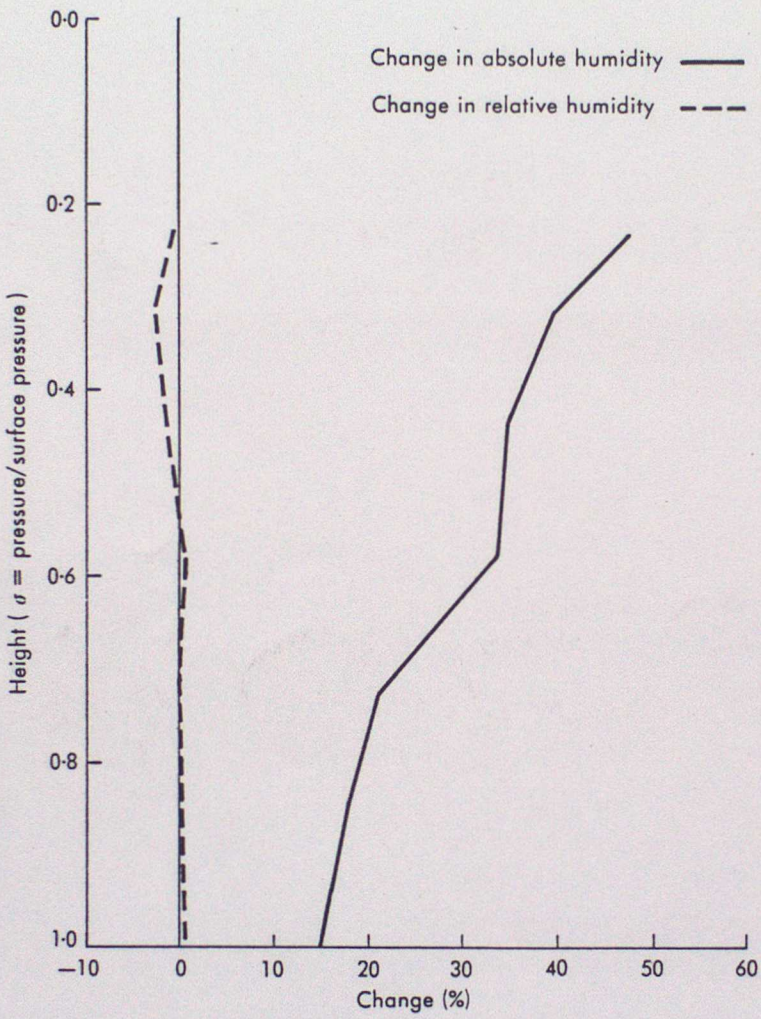
Change in absolute humidity
Change in relative humidity
0·0
0·2
0·4
0·6
0·8
1·0
Height (σ = pressure/surface pressure)
−10
0
10
20
30
40
50
60
Change (%)

Figure 1

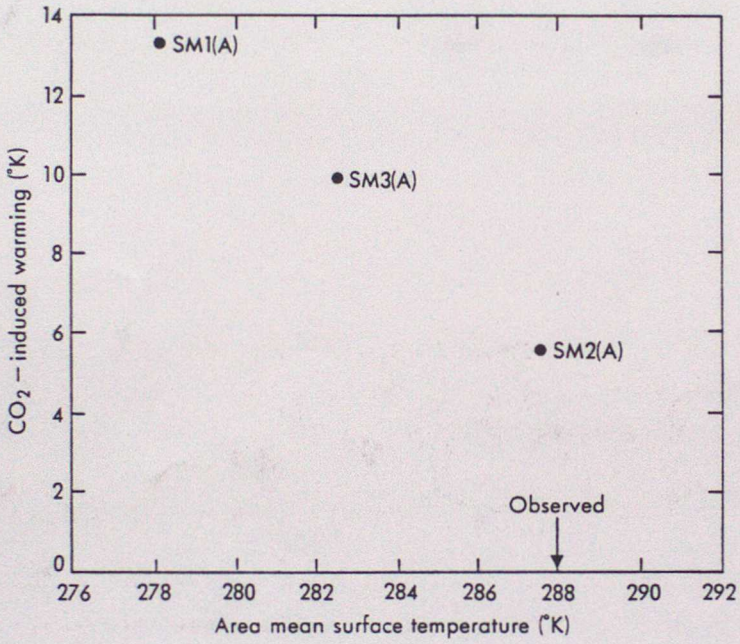
14
12
10
8
6
4
2
0
SM1(A)
SM3(A)
SM2(A)
Observed
CO_2 – induced warming (°K)
276
278
280
282
284
286
288
290
292
Area mean surface temperature (°K)

Figure 2

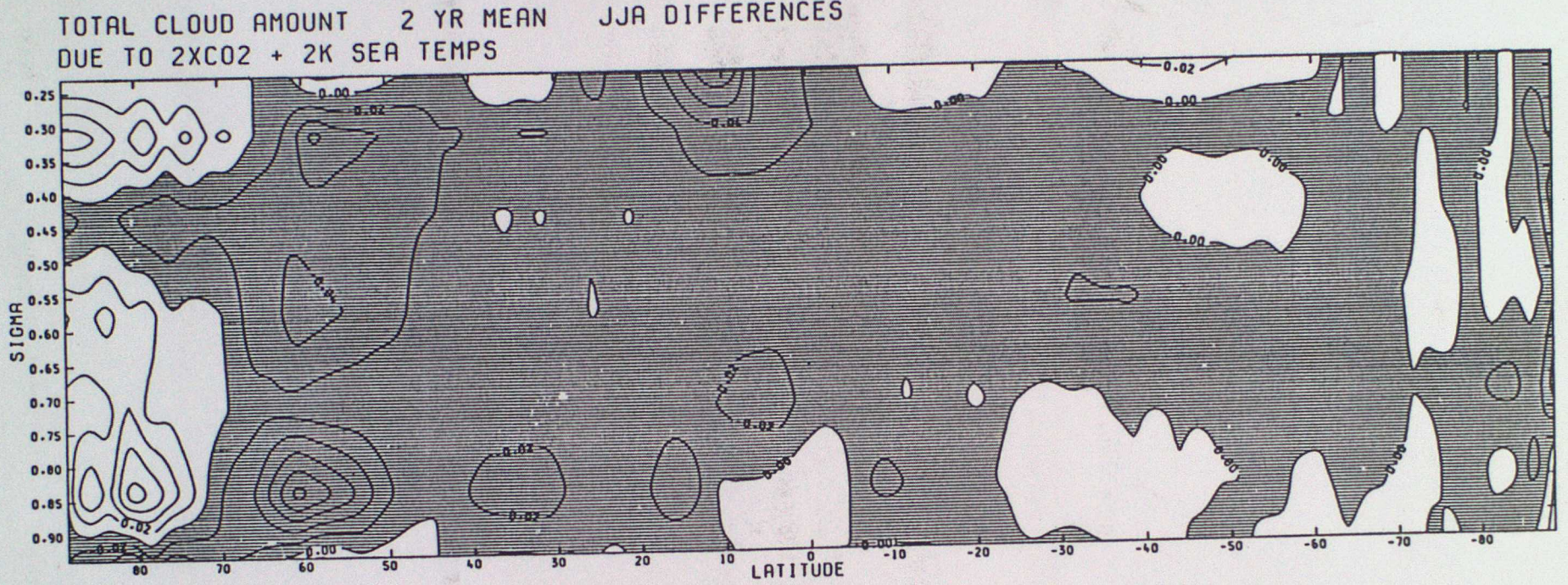
TOTAL CLOUD AMOUNT 2 YR MEAN JJA DIFFERENCES
DUE TO 2XCO2 + 2K SEA TEMPS
SIGMA
LATITUDE

Figure 3

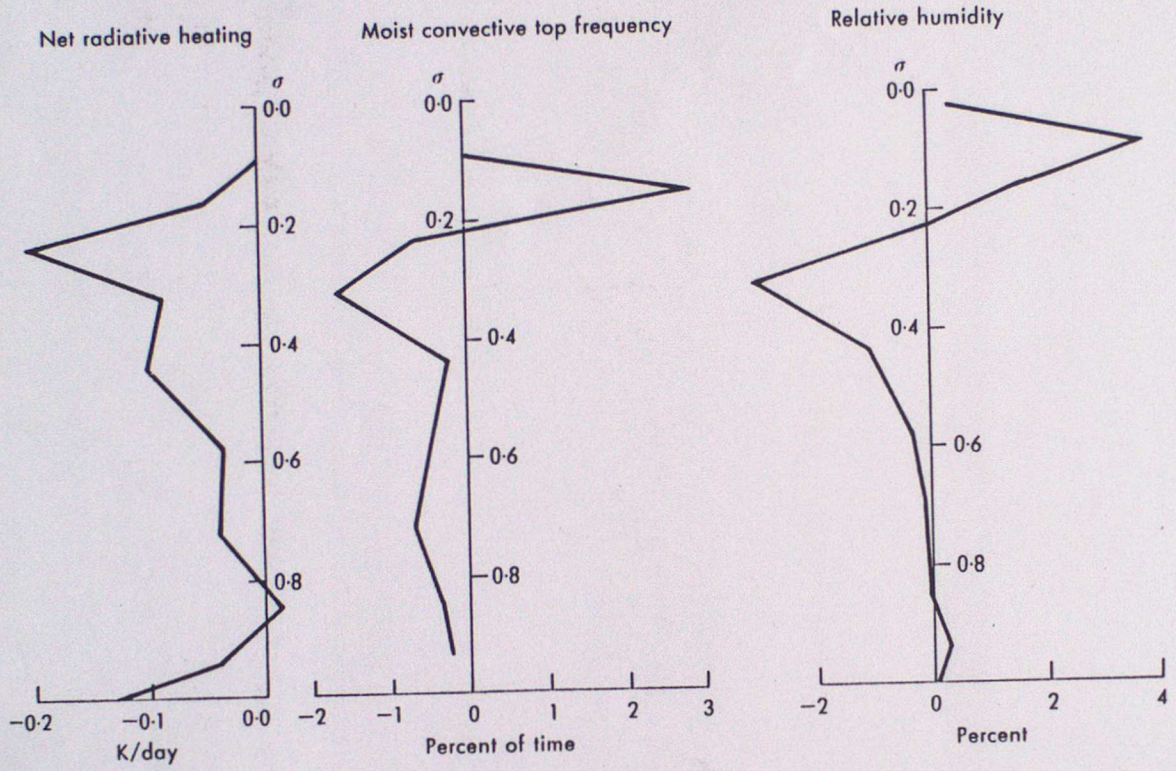
Net radiative heating
σ
0·0
0·2
0·4
0·6
0·8
−0·2
−0·1
0·0
K/day
Moist convective top frequency
σ
0·0
0·2
0·4
0·6
0·8
−2
−1
0
1
2
3
Percent of time
Relative humidity
σ
0·0
0·2
0·4
0·6
0·8
−2
0
2
4
Percent

Figure 4

SOIL MOISTURE CONTENT 2 YR MEAN JJA DIFFERENCES
DUE TO 2XCO2 + 2K SEA TEMPS

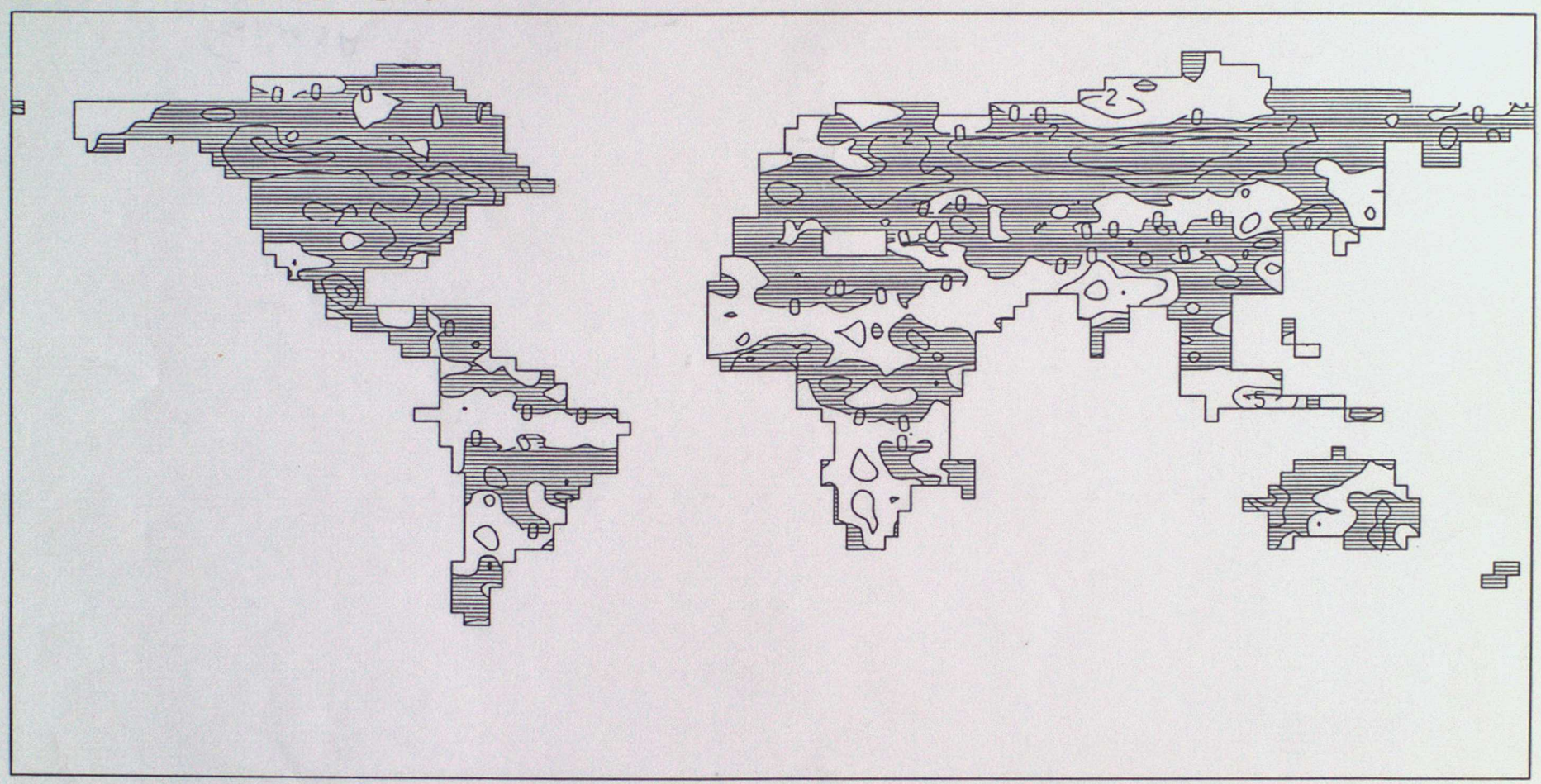

Figure 5a

TOTAL CLOUD AMOUNT 2 YR MEAN JJA DIFFERENCES
DUE TO 2XCO2 + 2K SEA TEMPS

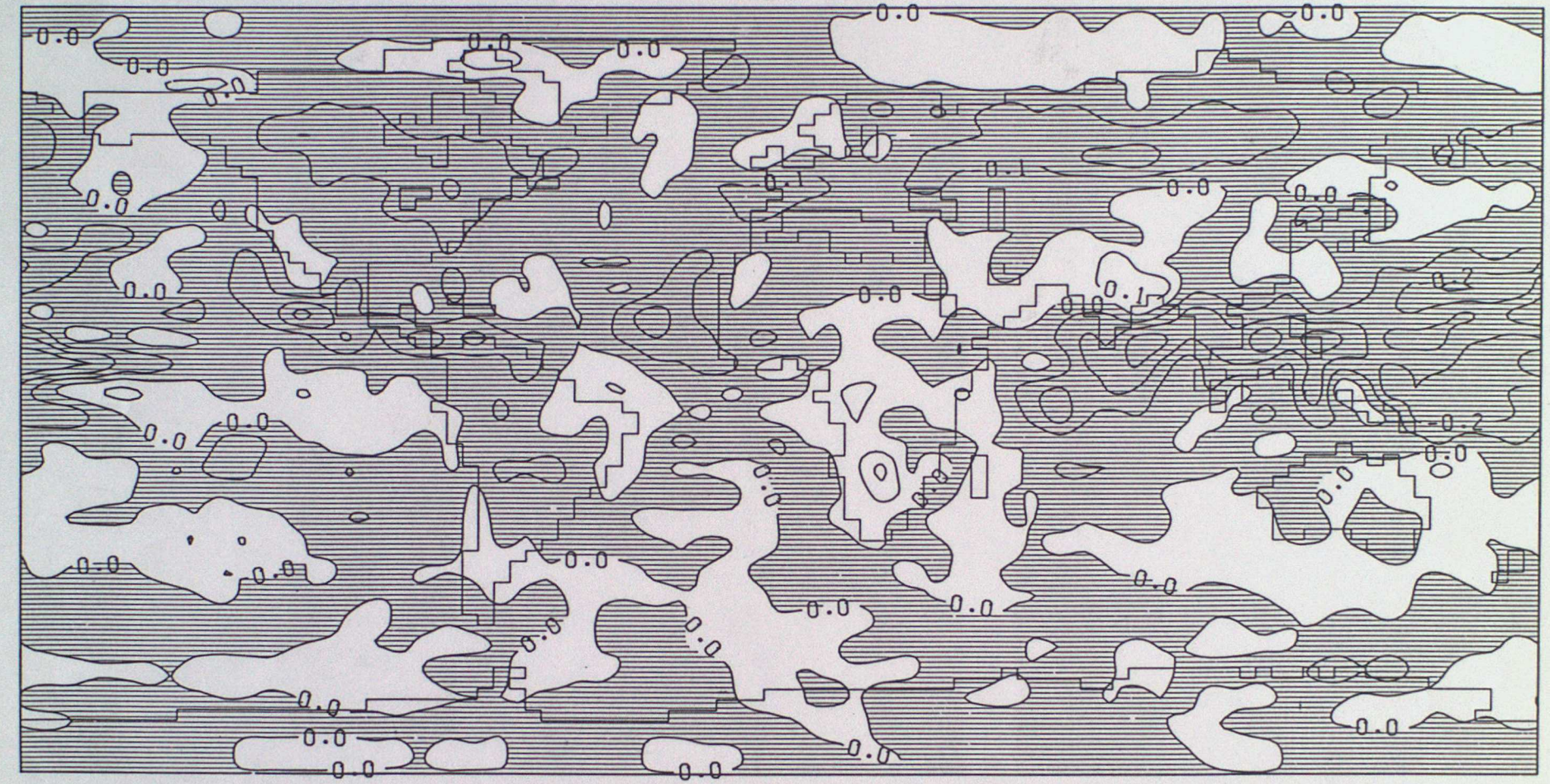

Figure 5b

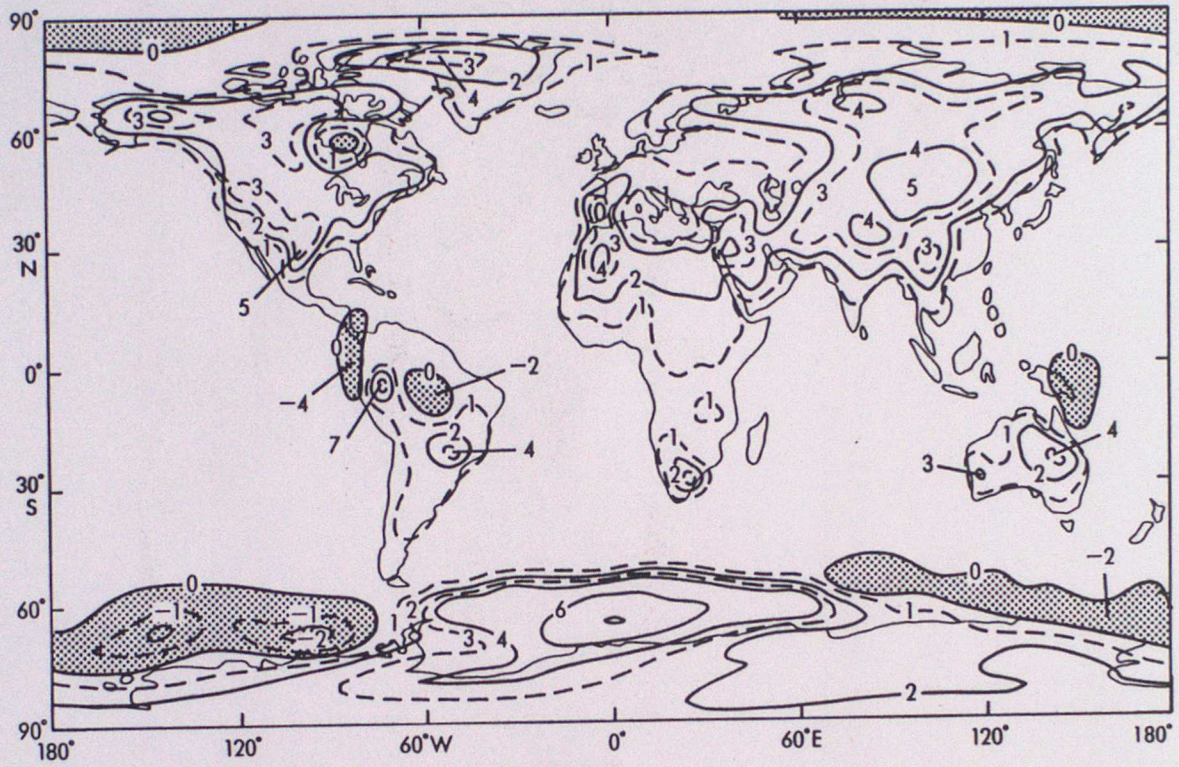

Figure 6